THE DETERMINATION OF EVAPORATION FROM LAND AND WATER SURFACES

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The problem of determining rate and amount of evaporation from land and water surfaces, or transpiration from vegetation, is rendered complex by the fact that the water becomes an invisible gas which rapidly mixes with the other gases of the atmosphere and is transported large distances and to great heights. A method for the accurate determination of water losses from lakes and reservoirs is of vital importance, but it is equally important to determine water losses from watersheds and from agricultural lands.

It is not possible to determine actual transpiration or evaporation from either free water surfaces or land surfaces by simply measuring the rate of loss of water from an exposed pan or atmometer. The amount of transpiration for any plant type is partly a function of leaf area, which varies seasonally with the growth cycle of the plant. Transpiration also varies from one plant species to another, depending on water requirements, the osmotic value in the leaves, and the number, nature, and size of the stomata. None of these factors is reflected by pan or atmometer measurements of evaporation.

Neither do these measurements indicate the evaporation from the soil. When the surface soil is moist the evaporation exceeds pan measurements because the soil with its minute irregularities presents a greater evaporating surface and because surface soil temperatures during the part of the day when most of the evaporation occurs are higher than water temperatures. When, however, the surface soil has become dry or partially dry, less evaporation occurs from the soil than from a pan. Even though the subsoil is moist, capillary action cannot supply the surface with water at a rate at all comparable to the evaporation from the surface of a body of water. Hence, water molecules can escape to the outer air only by a very slow diffusion process which takes place from the lower soil levels through the soil air, and results in a deceleration in the rate of moisture loss.

Since the rate of emission of water vapor from a land, water, or plant surface cannot be measured directly as can the rate of water accretion to the surface and since the measured rate from a pan or atmometer cannot give this rate of emission, the evaporation problem must be approached in a different manner.

Observations of evaporation from lakes, reservoirs, and pans have been used in the development of many empirical formulae in which evaporation is expressed as a function of various meteorological data such as temperature, relative humidity, barometric pressure, wind velocity, and solar radiation. Dalton (5) was the first to point out that evaporation is proportional to the difference between vapor pressure of the air at the water surface and that of the overlying air, although apparently he never expressed this relationship in mathematical terms.

In Rohwer's "Evaporation from Free Water Surfaces," published in 1931, (20) a number of evaporation formulae are presented and discussed. Most of these formulae contain the expression (e_s-e_d), e_s being the vapor pressure at the surface of the liquid and e_d the vapor pressure of the air, and a factor W expressing the influence of wind velocity. Rohwer's empirical formula is:

$$E=(0.44+0.118W)\ (e_s-e_d)$$

which, he says, is of the same form as Dalton's original formula:

$$E=C(e_s-e_d).$$

A different approach to the problem of determining evaporation from lakes and other bodies of water has been suggested by Ångström (2). He indicated that there must be a balance of energy between the insolation absorbed by a body of water and the energy exchanges due to radiation, convection, conduction, and the latent heat of evaporation, and that evaporation might be determined by measuring the necessary components of this equilibrium. Richardson (18, 19) and Cummings (4) have used this principle in determining evaporation losses but difficulties in obtaining various essential measurements, such as temperature gradients within the body of water and heat exchanges due to convection, prevent the determination of accurate short time measures of evaporation and prohibit the determination of moisture losses from land areas. Yamamoto (39) has made similar use of the conservation of energy principle in his studies of condensation of dew.

In recent years advances in aviation have stimulated the study of the structure of the lower atmosphere from which has come much essential information on turbulence and the intensity of turbulent mixing. In the free air there is a zone above the earth's surface in which the frictional effects of the ground are felt. This is the zone of turbulent mixing, or the turbulent layer. In this layer the mixing process is very important and depends on the shearing stresses associated with the roughness of the ground and with the wind structure, the thickness of the layer increasing as the value of these factors increases. The mixing process also depends upon the stability or density structure of the air, the mixing diminishing as stability increases. Beneath the turbulent layer there is a shallow layer usually measured in millimeters, called the laminar layer, in which air flow is laminar, where mixing proceeds by diffusion only. The vertical motion of the air in the turbulent layer tends to establish an adiabatic distribution of properties of the air, and thus to eliminate differences in moisture concentration. If moisture is neither added to nor withdrawn from the turbulent layer it quickly becomes uniform in moisture content throughout.

On the other hand, if water vapor is emitted from an evaporating surface it will be transported upward and scattered throughout the turbulent layer. Thus, as long as a stream of water vapor is flowing upward into the turbulent layer the moisture concentration will be highest at the base and will diminish upward, and a moisture gradient directed upwards will be established. Such a gradient could be maintained only as long as moisture continues to be added from below. When evaporation ceases the moisture will soon be distributed uniformly throughout the layer and the gradient will be destroyed. Similarly, if water vapor is abstracted from the base of the layer by condensation (dew or frost formation) the moisture concentration there will be reduced and as long as removal of moisture continues, a moisture gradient directed downwards will be maintained. The greater the intensity of turbulent mixing the greater will be the tendency toward the establishment of uniformity of moisture concentration, and the greater the evaporation or condensation required to maintain a constant gradient upward or downward. Similarly, the greater the moisture gradient which persists in an air column with a given rate of turbulent mixing, the greater will be the evaporation required to produce it.

From these facts it is evident that evaporation from any surface may be determined by taking into consideration the vertical distribution of moisture in the air and the intensity of turbulent mixing. The general form of an equation for evaporation based on these facts is $E=-A\,(dq/dh)$, in which A is the coefficient of turbulent mixing, or the Austausch coefficient of Willhelm Schmidt (24), and dq/dh is the rate of change of moisture concentration with respect to height above the evaporating surface. Observations required in order to measure evaporation are the moisture concentration at two levels within the turbulent layer, the heights of the two observations, so that moisture gradients may be obtained, and measurements of wind velocity at two or more levels to give the intensity of turbulent mixing. The moisture concentration in the air is measured as specific humidity. Within the height considered where air pressure is essentially constant the specific humidity is proportional to the vapor pressure.

Jeffreys (9) and Giblett (7) were among the earliest investigators who studied the problem of evaporation in terms of atmospheric turbulence. However, because of various nonrigorous assumptions regarding the nature of the Austausch coefficient, their results are limited and can be applied only to the evaporation from bodies of water.[1] Sutton (26), making use of Taylor's (32) researches, has extended Jeffrey's analysis by assuming that the Austausch coefficient varies with height. His theoretical work was found to be in very good agreement with experimental evaporation measurements for variously shaped areas. Schmidt (25) has made use of a formula devised by Ertel (6) for calculating the Austausch coefficient and claims to have measured the actual evaporation from a meadow.

During the last two decades great progress in the theoretical treatment of the mechanism of turbulent interchange has been made by various aerodynamicists, including Taylor (30, 31, 32, 33, 34), Prandtl (16, 17), von Kármán (10, 11, 12, 13), and others. In 1935 von Kármán (12) published an excellent summary of the problems associated with the various theories of turbulence. Rossby (21, 22) has made use of von Kármán's postulate of dynamic similarity and the concept of mixing length in his important researches on atmospheric turbulence. Sverdrup (27), working along lines explored by Rossby, has added much to our knowledge of the variation of the coefficient of eddy viscosity under changing conditions of thermal stability.

Although the theoretical treatment of problems of turbulence is still in the initial stage, sufficient has been accomplished to provide an understanding of the mechanism of evaporation and to permit the development of a formula by means of which evaporation from land and water surfaces alike may be determined.

A difficulty in the use of turbulent theory in the determination of evaporation is the fact that the laws of moisture transfer are different in the laminar and turbulent layers, being linear in the former and nonlinear in the latter. Leighly (14) avoided the turbulent layer entirely and confined his study to the laminar or boundary layer where the specific humidity gradient is linear and where the coefficient of diffusion can be used. Since the boundary layer is very thin if the air is in motion, Leighly's method would be useful only for determining evaporation from small areas (he thinks that the limit is a diameter not much greater than 4 feet). In fact, the making of moisture determinations at two levels in a layer as thin as the boundary layer would require a type of instrumentation not now available. With existing instruments it is quite impossible to recognize vapor pressure differences within a distance of a few millimeters. Thus, the use of Leighly's method must await the development of new instruments.

The problem of the difference in character of the moisture gradient and the process of moisture transfer in the boundary and turbulent layers confronted Sverdrup (28, 29) in his study of evaporation over the oceans. With the meager data at his disposal, consisting only of wind velocity and moisture concentration at one level in the turbulent layer, and the temperature and salinity of the ocean water, he was forced to consider the gradients in both the laminar and turbulent layers. Although this was accomplished with great ingenuity, the final computations can be considered only as rough indications of oceanic evaporation. Millar (15), in his study of evaporation from the Great Lakes, has approached the problem in much the same manner that Sverdrup did for the oceans, and his results are subject to the same limitations.

At this point it is possible to consider the physical limitations of the various empirical formulae based on Dalton's law, which have been presented in an earlier paragraph. These formulae are developed for determining evaporation from free water surfaces and make use of observations of the temperature of the water surface (from which the vapor pressure of the air in direct contact with the water is determined) and the vapor pressure and wind velocity at some level in the turbulent layer. Because of the previously mentioned difference in character of the moisture gradients and the processes of moisture transfer in the boundary and turbulent layers, it is impossible to derive simple empirical constants, to be used in connection with the difference between the vapor pressure of the water surface and the overlying air (e_w-e_d) and a wind factor, which will give more than a statistical approximation of the evaporation from free water surfaces. Using Sverdrup's method of integrating through the boundary layer, more satisfactory results, at least for large water bodies, could be expected. Under no circumstances could such observations be used for determining evaporation from land or transpiration from vegetation. Those formulae employing the so-called vapor pressure deficit (1), the difference between the saturation vapor pressure and the actual vapor pressure at some level in the air (E_d-e_d), are unrelated to the physical principles of evaporation, and cannot be expected to yield satisfactory results.

As an outgrowth of the researches on turbulence by von Kármán and Rossby and through discussion with them, a formula for determining evaporation from observations of moisture concentration and wind velocity in the turbulent layer has been devised and is here presented. The derivation of the formula follows directly from an expression for the Austausch coefficient obtained from the concepts of the mixing length and the shearing stress as developed by Prandtl and von Kármán. The derivation is straightforward and can readily be repeated by referring to the works of Rossby (21, 22) and Sverdrup (27). The formula itself is:

$$E=\frac{K_0^2\rho(q_1-q_2)u_2}{\log_e\frac{h_2}{h_1}\,\log_e\frac{h_2}{Z_0}}$$

in which K_0=von Kármán's coefficient.
ρ=density of the air.
q_1=moisture concentration at the lower level,
q_2=moisture concentration at the upper level,
u_2=wind velocity at the upper level,
h_2=height of the upper instruments,
h_1=height of the lower instruments,
and Z_0=roughness coefficient.

[1] A brief critical discussion of the studies of Jeffreys and Giblett may be found in Brunt (3), pp. 264–267.

The roughness coefficient is determined from observations of wind velocity at two levels by means of the following formula:

$$\log_e Z_0 = \frac{u_2 \log_e h_1 - u_1 \log_e h_2}{u_2 - u_1}$$

Since the intensity of turbulent mixing is dependent on the wind velocity and the roughness coefficient and since the latter can be determined from wind velocities at two levels, it is possible to simplify the evaporation formula to the following:

$$E = \frac{K^2{}_0\rho(q_1-q_2)\ (u_2-u_1)}{\left(\log_e \frac{h_2}{h_1}\right)^2}$$

In the above formulae all values are in *CGS* units. The formula giving evaporation in inches per hour for an installation where the upper observations are taken 28.6 feet and the lower ones 2 feet above the ground is:

$$E = \frac{0.0274\ P(q_1-q_2)\ (u_2-u_1)}{T+459.4}$$

where P is pressure in inches Hg

q_1 and q_2 are in grams per kilogram
u is wind velocity in miles per hour
and T is temperature in °F.

Expressed in terms of vapor pressure the formula is simplified further, and becomes

$$E = \frac{17.1(e_1-e_2)\ (u_2-u_1)}{T+459.4}$$

in which e_1 and e_2 are the vapor pressures in inches of Hg at the lower and upper levels, respectively.

The formulae as presented are rigorously correct for an adiabatic atmosphere but in cases of thermal stability a correction must be applied for a precise evaluation of evaporation. It is now known that when the temperature and wind structure of the air attain certain critical values (**23**) the effect of turbulent mixing can be completely suppressed and the moisture transfer must then proceed by diffusion alone. Under conditions of light winds and temperature inversions, therefore, the values for evaporation or condensation, as the case may be, determined by the use of the above formulae, are too large. This introduces no serious error in computations of evaporation for periods of a day or longer as the amounts of moisture evaporated under conditions of stability are negligible compared with those amounts evaporated during adiabatic conditions. For the determination of the amount of dew or frost deposition the adiabatic formula becomes inadequate when atmospheric stability is such as to bring about a marked suppression of turbulence. A correction for the stability factor devised along the lines suggested by Rossby (**21, 22**) and Sverdrup (**27**) is being tested at the present time. However, since the amounts of condensation are quite small, again no serious error is involved in the consideration of moisture losses from watersheds.

Two experimental evaporation stations have been established; one at the Muskingum Climatic Research Center in Ohio and the other on Arlington Farm in Virginia. At the Arlington installation the instruments are located in a hay field on the relatively flat top of a gentle ridge. The hay had been cut shortly before the installation was made, and being composed of perennial grasses, continued to grow and remained more or less green throughout the winter. In each case a tower has been erected in such a manner that an instrument shelter large enough to house a hygrothermograph may be raised to a height of 28 feet and lowered at will (fig. 1). Another shelter accommodates a second hygrothermograph 2 feet from the ground. Records from these hygrothermographs, together with records from a barograph, are the only requirements for determining the specific humidity at the two levels. Recording anemometers at these levels supply the other information required in the evaporation formula (fig. 2). At the present time three hygrothermographs and four anemometers, mounted at different elevations on a tower, are being operated at Arlington to supply information for testing the validity of an expression derived to correct the adiabatic formula for cases of atmospheric stability.

Despite the fact that the hair hygrometer as an instrument for measuring relative humidity has many defects, the chief of which are the uncertainty of registration of very high and very low values, the hygrothermograph is found to be the most satisfactory existing instrument for our purpose.[2] In order to be certain of the results, six instruments have been used on each installation. All six instruments were first carefully calibrated in a moisture chamber where, by means of saturated solutions of various salts,[3] constant humidities could be maintained at various levels, and a separate calibration chart was prepared for each instrument. They were then grouped in pairs and were used in succession on the tower. Before two instruments were used they were given a background run side by side in a Weather Bureau shelter. After use they were given another background run and were again calibrated in the moisture chamber. Three datum pens were added to each instrument to permit correction for shrinking and stretching of the chart paper. Recently nonhygroscopic metal charts have been used in place of the moisture-sensitive paper charts and the index pens are no longer needed. Safe-guarded in this manner the results are felt to be reasonably reliable.

The first usable records of moisture concentration were obtained on October 15, 1938. Continuous daily observations on the moisture gradients have been made since the latter part of November. From these records it has been possible to get some indication of the diurnal changes in moisture concentration and gradient for certain air mass types and of the magnitude of the moisture gradients. Records of wind velocity at the two levels have been available only since November 23, 1938, and not until December 10, 1938, were actual determinations of evaporation begun.

Figure 3 shows the march of moisture concentration at the two levels for two days in early autumn during which time a modified Polar Continental air mass remained over the area. The sky was clear and moderate temperatures prevailed. The air mass was relatively stagnant with light and variable winds. Significant radiational cooling occurred throughout the night and a moderate ground fog condition limiting visibilities to 1 to 2 miles was observed at the site of the experimental station.

During the first night the moisture concentration of the air dropped to a minimum of 5.4 and 5.8 grams in the lower and upper levels due to abstraction of moisture by condensation at the ground. As soon as the ground began to

[2] An automatic dewpoint recorder has been devised by the authors and a laboratory model has been built by Dr. L. R. Hafstad of the Carnegie Institution of Washington and J. C. Owen of our staff. If satisfactory field instruments can be made they should be more accurate than any type of instrument now available.

[3] The salts used and the relative humidities which they maintain at room temperatures are as follows: KNO_3, 95 percent; $NaCl$, 75 percent; $Mg(NO_3)_2.6H_2O$, 52 percent; $MgCl_2.6H_2O$, 32 percent; $LiCl$, 16 percent.

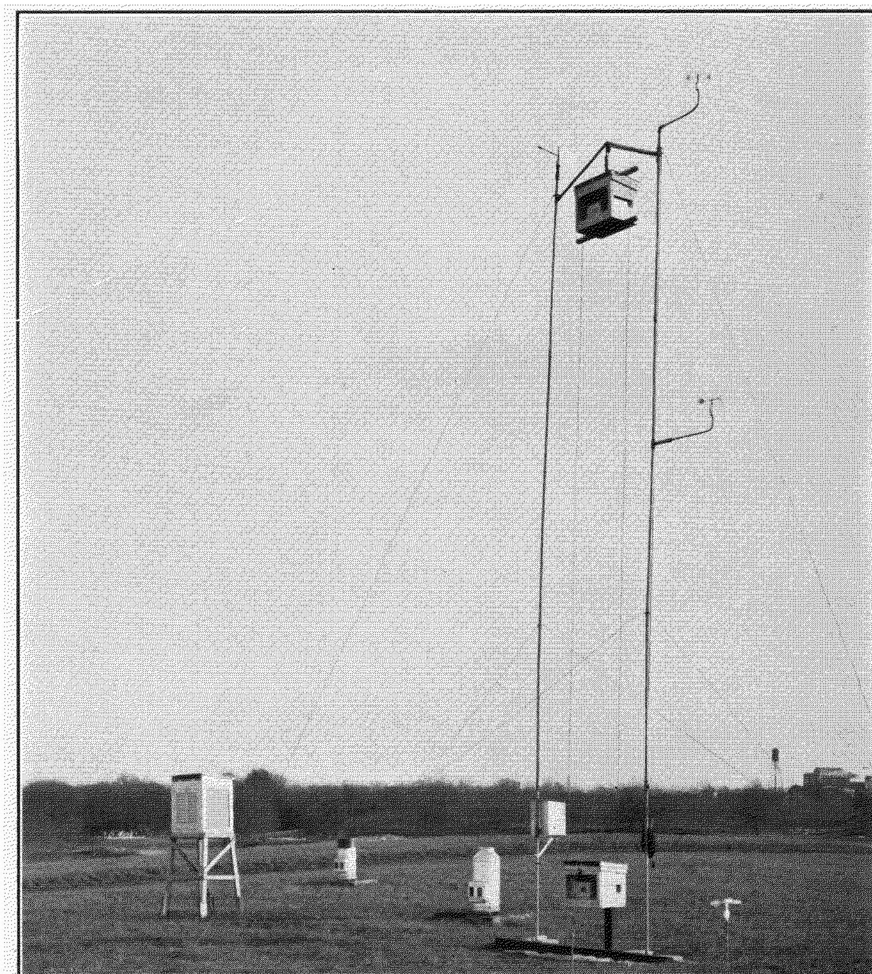

FIGURE 1.—Evaporation instrumental installation at Arlington Farm, Va.

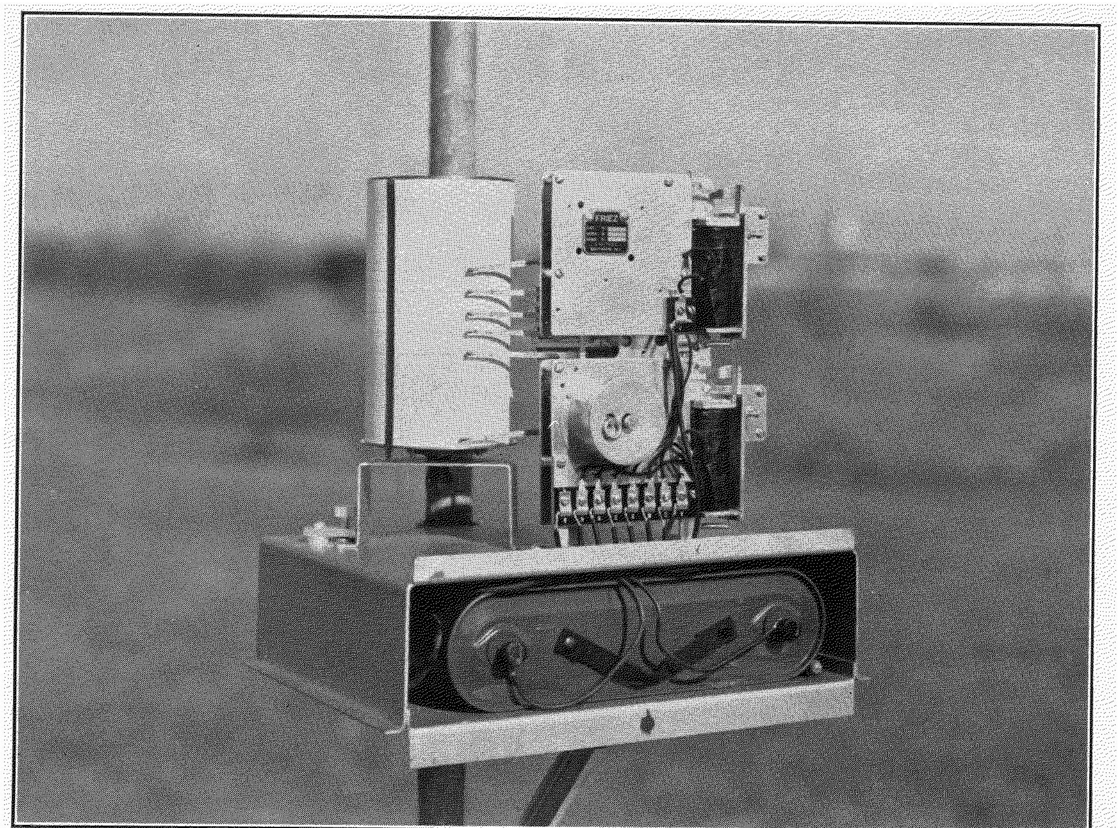

FIGURE 2.—Recorder for wind velocity at two levels and wind direction at one level, designed especially for evaporation studies. A similar instrument giving a synchronized record of wind velocity at four levels and wind direction at one level has recently been installed.

TABLE 1.—*Data for the computation of evaporation at Arlington Farm, Va., for the 24-hour period Jan. 27, 1939*

	Hour ending—A. m.											
	1	2	3	4	5	6	7	8	9	10	11	Noon
Upper observations:												
Relative humidity	66.2	70.8	75.2	80.1	81.5	75.1	52.0	45.4	43.5	45.2	41.0	39.4
Temperature	20.8	20.3	20.1	18.9	18.7	19.3	22.8	23.3	24.0	25.1	26.5	28.5
Vapor pressure	.071	.074	.078	.079	.079	.075	.061	.055	.054	.059	.057	.061
Wind	5.0	3.8	4.0	1.8	1.5	3.5	9.0	5.8	13.5	12.8	19.2	12.5
Lower observations:												
Relative humidity	74.6	76.7	80.9	84.9	87.9	84.2	58.8	52.5	51.4	50.0	45.5	42.0
Temperature	19.4	18.5	17.5	16.5	16.0	16.5	20.5	21.0	22.4	24.9	26.9	29.4
Vapor pressure	.075	.074	.074	.074	.075	.073	.062	.057	.060	.065	.065	.067
Wind	3.2	2.0	1.8	1.0	.2	1.5	5.8	3.5	7.5	8.2	12.8	8.5
Difference:												
Vapor pressure	.004	.000	−.004	−.005	−.004	−.002	.001	.002	.006	.006	.008	.006
Wind	1.8	1.8	2.2	.8	1.3	2.0	3.2	2.3	6.0	4.6	6.4	4.0
Evaporation [1]	.0003	.0000	−.0003	−.0001	−.0002	−.0001	.0001	.0002	.0013	.0010	.0018	.0008
	Hour ending—P. m.											
	1	2	3	4	5	6	7	8	9	10	11	Midnight
Upper observations:												
Relative humidity	36.5	34.2	31.0	32.0	34.1	35.3	35.5	35.5	37.6	41.0	43.0	47.5
Temperature	30.5	32.5	34.0	34.0	34.0	33.0	32.3	31.7	31.0	30.0	28.3	26.5
Vapor pressure	.061	.063	.061	.062	.067	.066	.065	.063	.065	.068	.065	.067
Wind	13.5	15.2	14.8	12.2	14.5	11.2	8.0	7.2	5.0	3.5	2.8	4.2
Lower observations:												
Relative humidity	38.0	34.2	32.0	34.5	36.9	39.3	41.5	41.1	43.1	50.0	64.0	71.4
Temperature	31.9	33.9	34.9	33.6	32.9	31.4	30.4	29.4	29.4	26.9	22.9	21.4
Vapor pressure	.068	.067	.065	.066	.069	.069	.070	.066	.069	.071	.076	.079
Wind	9.2	10.5	9.8	9.8	9.0	8.0	5.0	4.2	2.8	1.8	.2	1.2
Difference:												
Vapor pressure	.007	.004	.004	.004	.002	.003	.005	.003	.004	.003	.011	.012
Wind	4.3	4.7	5.0	2.2	5.5	3.2	3.0	3.0	2.2	1.7	2.6	3.0
Evaporation [1]	.0011	.0007	.0007	.0003	.0004	.0003	.0005	.0003	.0003	.0002	.0010	.0013

[1] Negative evaporation is condensation. Totals for the period. Evaporation 0.0126 inches, Condensation 0.0007 inches.

warm up after sunrise and evaporation commenced, moisture concentration rose rapidly, beginning at the lower level and producing a reversal in the gradient. Shortly before noon thermal convection had developed to a point where the moisture was carried to the upper air faster than it was received from the ground and the concentration at both levels diminished about 0.8 grams. In late afternoon, with a diminution of convection, the evaporation again exceeded the rate of transport aloft and the moisture concentration rose to the high point of the day. Between 6 and 7 p. m. the cooling of the ground surface began, and soon the concentration at both levels was the same. With continued cooling condensation was initiated, and as moisture was abstracted from the lower level the gradient was reversed, and the concentration at both levels diminished throughout the night. In the morning there was a heavy dew deposit.

The same sequence of conditions was repeated the next day; first an upward shift of gradient followed by a rapid increase of moisture concentration as evaporation proceeded; then a diminution of concentration with a continued upward-directed gradient as thermal convection carried the moisture to the drier levels aloft; then as convection subsided a rapid increase in concentration; and finally, with the onset of ground radiation at the end of the day, condensation, a gradient directed downward, and a rapid diminution of moisture.

A few days later, with the passage of a well-marked front the transitional Polar Continental air mass was displaced by a fresh polar air mass. The moisture concentration dropped within the course of a few hours from about 10.5 to 4.0 grams. The moisture gradient pattern for the 2 days following is shown in figure 4. Except for scattered cirrus and a few scattered cumulus clouds during the afternoon the sky was clear throughout the 2-day period. For nearly the entire first day the wind direction was predominantly from the west and west-northwest with velocities ranging from 5 to 15 miles per hour. The moisture gradient, although small, was directed upward for approximately 28 hours after the invasion of the polar air mass. The moisture concentration remained almost stationary at about 4 grams until approximately 3 p. m., when a moderate amount of convection forced a small decline. It is of interest to note that scattered cumulus clouds were observed between 3 and 5 p. m., which coincided with the observed decline in moisture concentration. During the evening evaporation continued and the concentration rose more than a gram. Not until after 10 p. m. did the gradient turn downward and condensation commence.

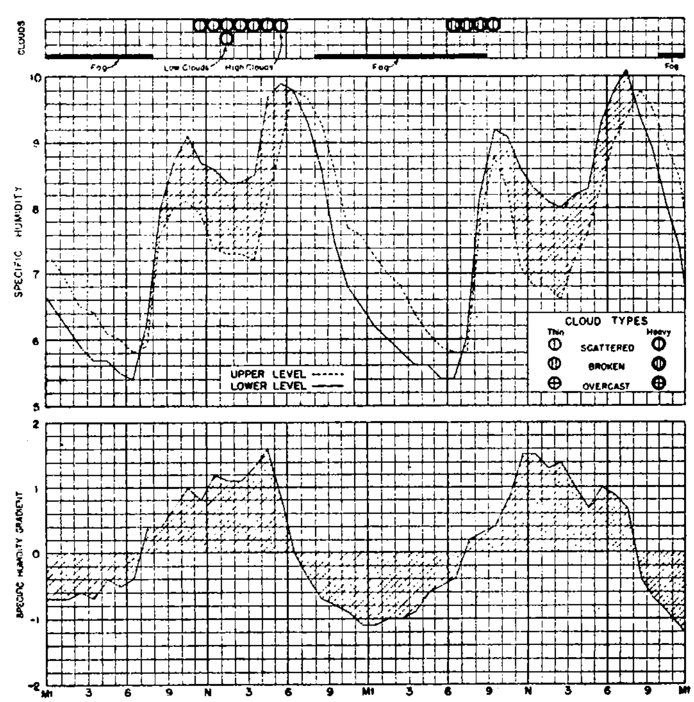

FIGURE 3.—March of moisture concentration at two levels, and of the moisture gradient, for October 15 and 16, 1938. Arlington Farm.

At this time the wind had subsided, becoming variable with velocities generally less than 2 to 3 miles per hour. Thereafter during the night, with moisture abstraction going on at the ground, the moisture concentration slowly declined. The gradient pattern of the second day resembled those of the 2 earlier days shown in figure 3,

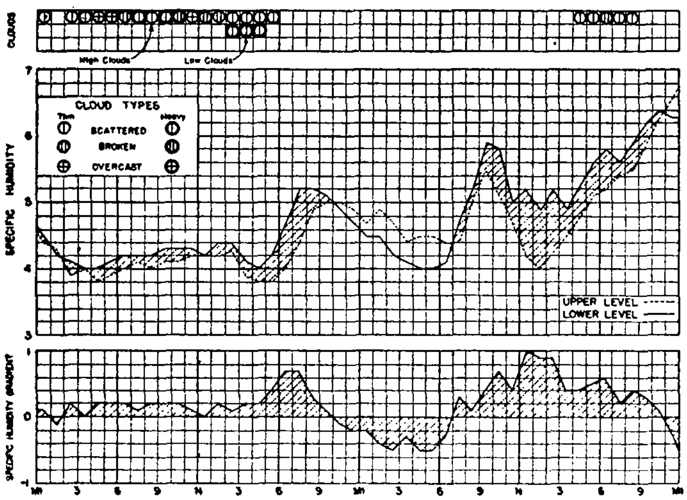

FIGURE 4.—March of moisture concentration at two levels, and of the moisture gradient, for October 21 and 22, 1938. Arlington Farm.

although the rate of evaporation was considerably reduced. The important difference in pattern is that the moisture gradient was directed upward until 11 p. m., with the moisture concentration continuing to rise until that time.

For the cases already discussed, wind velocity gradients

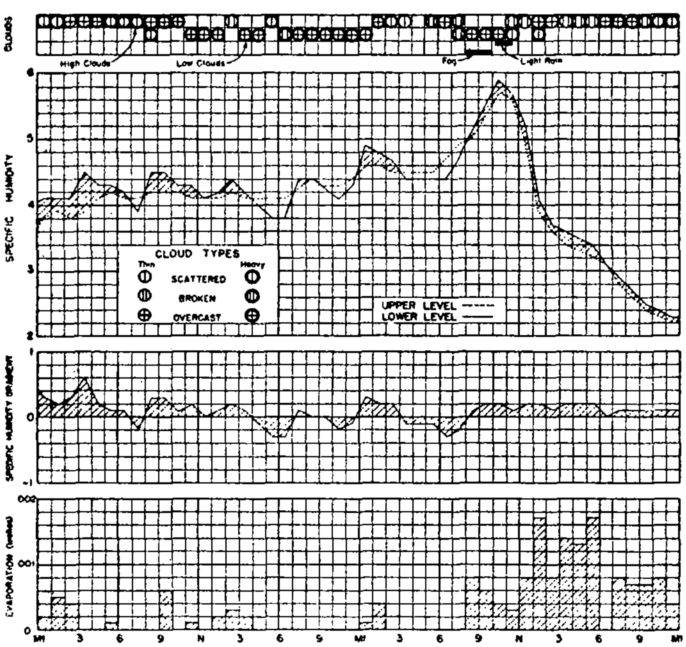

FIGURE 5.—Moisture concentration at two levels, moisture gradient and hourly evaporation for December 11 and 12, 1938. Arlington Farm.

were not available, consequently it was not possible to compute the actual evaporation.

Complete data for calculating the evaporation from the ground have been obtained with occasional unavoidable interruptions since November 23, 1938. Figures 5 and 6 show the moisture gradient pattern and the hourly rates of evaporation for December 11 to 14, during which time a Polar Continental air mass displaced a southerly current of well-modified polar air. On December 11, the synoptic weather chart indicated the presence of a well-modified returning flow of polar air over the eastern United States. The sky conditions as observed at the Washington airport adjoining the Arlington Farm, were quite variable from hour to hour. The thin, broken to overcast, altostratus clouds which predominated during the early morning and forenoon thickened to a nearly continuous overcast with a base at 6,000 to 7,000 feet during the afternoon and evening. During this time the specific humidity gradient was directed upward. Between 5 and 7 p. m. the clouds became scattered and broken and radiational cooling of the ground hastened the usual reversal of the specific humidity gradient. As the overcast thickened again, the gradient shifted upwards temporarily for an hour. Thereafter the gradient was directed downward except for a period between midnight and 3 a. m. of December 12. The upward gradient at this time was probably due to a

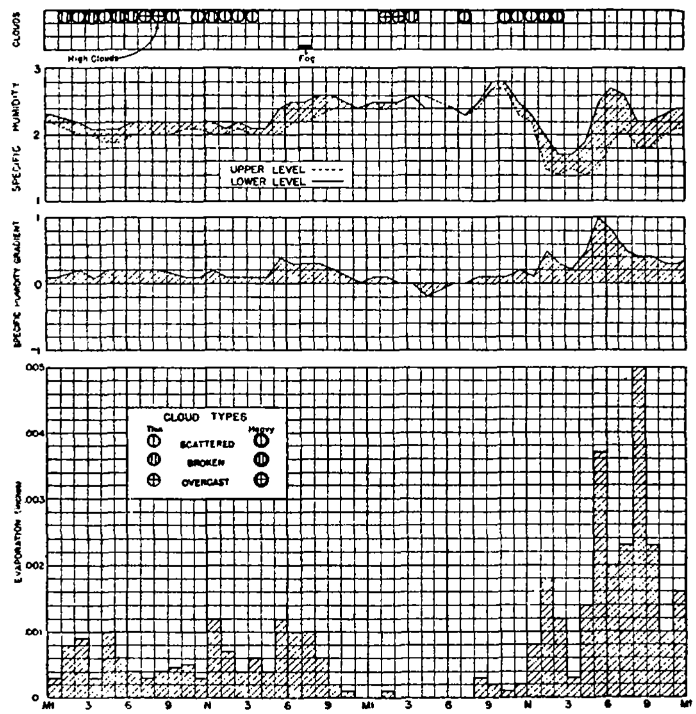

FIGURE 6.—Moisture concentration at two levels, moisture gradient and hourly evaporation for December 13 and 14, 1938. Arlington Farm.

slight increase in surface temperature caused by the downward transport of heat by turbulence induced by the increased wind velocities which prevailed during this period.

During the early morning of December 12, the clouds were, in general, high and scattered and calm or extremely light winds prevailed. The moisture gradient continued downward and a light dew formation resulted. At 9 a. m. the sky became overcast with upper altostratus and lower strato-cumulus decks heralding the approach of a cold front. Occasional sprinkles were noted during the hour ending at 9 a. m., and light rain was observed between 10:10 and 11:25 a. m. The amount of rainfall, however, was less than 0.01 inch. The front, which marked the boundary of Polar Continental air, passed the station at approximately 11 a. m. and was associated with a wind shift to the northwest and a dissipation of the strato-cumulus deck. During the ensuing 30 hours the wind remained in the northwest with velocities generally between 8 and 15 miles per hour. The specific humidity gradient was directed upward for 2 hours prior to the frontal passage and continued to be directed upward until the morning of the 14th. The gradient shifted downward at this time and a heavy frost ensued.

It is to be noted that on the 13th there was no drop in specific humidity resulting from large scale thermal or convective turbulence such as was illustrated in figure 3. This is explained by the fact that convective turbulence was limited to the thick layer of polar air, within which mixing by mechanical turbulence had already established complete uniformity of moisture concentration, thus precluding the possibility of replacement of the air at the level of the observations by drier air from aloft.

On December 14, however, the wind velocities in the polar air were light and variable and the thickness of the mechanically turbulent layer was much less than that on the 13th. From 11 a. m. to 3 p. m. the characteristic convective turbulence was manifested by a steady drop in the specific humidities in spite of the fact that evaporation was continuously adding moisture to the atmosphere. A secondary decrease in the specific humidities beginning at 8 p. m. is also due to mixing with drier air aloft. The wind velocities increased at this time, shifting from south at 3 miles per hour to west-northwest at 15 miles per hour, causing an increase in surface temperature from 38° to 42°.

The water loss from the land surface at Arlington Farm, combining evaporation from the soil, evaporation of dew and frost, and transpiration, for the 4 days from December 11 to December 14 was 0.0026 inches, 0.0137 inches, 0.0132 inches and 0.0239 inches respectively.

The station was operated without interruption from November 23, 1938 until February 7, 1939 and hourly evaporation has been determined for this entire period. Figure 7 shows the daily evaporation and condensation for the month of January. In table 1 the type of information obtained and hourly evaporation values for January 27, 1939 are presented merely for illustrative purposes.

During the first 4 days in January the ground surface was dry and the moisture was apparently removed chiefly by transpiration (**37**). Variable high scattered clouds and light variable winds prevailed during this period and appreciable amounts of dew condensed during the early morning hours. On January 5 a low stratus overcast prevailed throughout the greater portion of the day. Intermittent sprinkles and occasional light rain totaling 0.07 inches occurred during the 24-hour period. In spite of the cloudy conditions and the precipitation a measurable moisture gradient directed upward was detected and 0.0103 inches of evaporation occurred. The following day, January 6, the skies were relatively clear, moderate wind velocities were observed, and the evaporation was greatly increased, amounting to 0.0544 inches. On the 7th, 8th, and 9th variable high clouds to clear skies prevailed but the daily evaporation totals were, as was to be expected, less than that of the 6th when the ground surface was moistened by the previous day's rainfall. On the 8th the evaporation was slightly greater than on the 7th or 9th and may be explained by the moderate wind velocities, occasionally reaching 22 miles per hour. Light rain amounting to 0.07 inches fell between 7 a. m. and 9:30 a. m. on January 10. In spite of the moisture on the ground and a moderate south-southwest wind that varied between 8 and 12 miles per hour, the evaporation amounted to only 0.0276 inches. The suppression of evaporation on this day may be explained by the diminution of the moisture gradient due to the invasion of more moist air from the south, causing an increase in specific humidity from 3.5 grams to approximately 7 grams. On the following day, there was only slightly more evaporation. However, by the morning of the 12th fresh polar air had invaded the area and the specific humidity dropped to 2.2 grams. The wind blew prevailingly from the northwest with a velocity ranging between 5 and 15 miles per hour, and in spite of a high overcast the evaporation attained a value of 0.0535 inches. The increased evaporation into polar air masses, which has been found consistently in our observations, was anticipated on the basis of the cycle of air masses (**8, 35**).

The high alto-stratus overcast that was present throughout the 12th steadily lowered and thickened, heralding the approach of a warm front storm. On the 13th a mixed precipitation of freezing rain, sleet and snow fell throughout the greater part of the day. On the 16th a light snow fell between 6 a. m. and 11 a. m. On the 18th snow began falling shortly after midnight and continued throughout the day until 10 p. m. A light snow, associated with an upper cold front, occurred on the 20th between

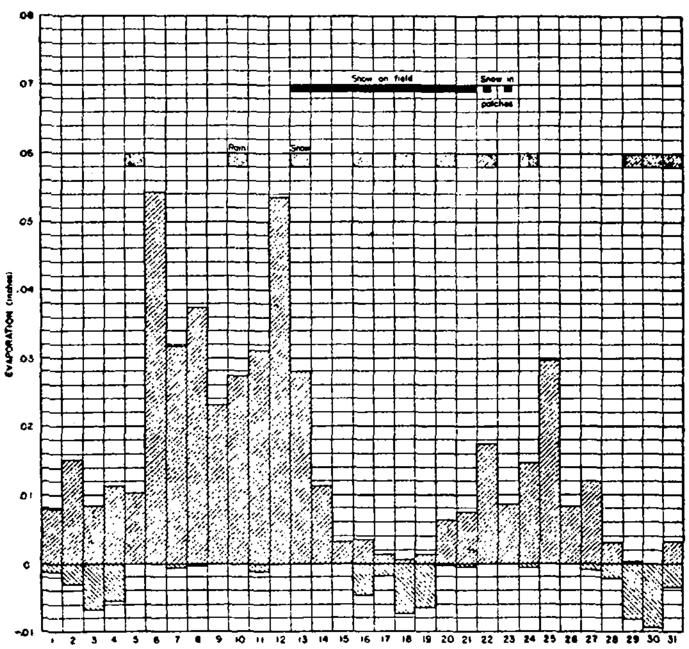

FIGURE 7.—Daily totals of evaporation and condensation for the month of January, 1939. Arlington Farm.

5 p. m. and 9 p. m. Thus from the 13th through the 20th the site of the evaporation station was covered with snow ranging in thickness from 1 to 6 inches.

During this period the blanket of snow was fundamentally responsible for the suppression of evaporation. A snow cover may suppress evaporation in a number of different ways, although it is by no means intended to imply that the moisture losses by evaporation from a snow field may not be large when suitable atmospheric conditions prevail. Because of the difference in the physical characteristics of a snow cover and a grass cover there must obviously be a difference in the roughness factor characterizing the type of surface and a consequent difference in the intensity of turbulent mixing.

Rossby and Montgomery (**22**, p. 10) using observations published by Shaw have determined the roughness factor for an open grassland. Sverdrup (**27**, p. 44) has made a similar determination for a snow field from observations taken in West Spitzbergen. Using these determinations of roughness it is found that under otherwise similar conditions the evaporation from the meadow would exceed by approximately 50 percent that from the snow field.

Similar calculations for Arlington Farm showed a reduction of the roughness factor during the period when the ground was snow covered that was in excellent agreement with this figure.

An important factor governing the rate of evaporation from the snow is the temperature of the snow surface. It is well known that a snow surface radiates energy as a black body (38) and on a clear calm night, excessive inversions of temperature may be established in the air next to the snow. During the daylight hours much of the incident radiation is reflected and the snow surface temperatures may remain well below the dewpoint temperatures of the air. In any event the temperature of the snow surface must remain at the freezing temperature until all of the snow is melted. Comparing a grass-covered field with a snow field, the former will also radiate energy like a black-body but during the daylight hours most of the incident energy will be absorbed and the surface temperatures may be greatly increased, resulting in increased transpiration of the plants. Transpiration may be regarded as the most effective mechanism whereby moisture is returned to the atmosphere from land areas (**37**). A snow cover obviously precludes the operation of the process of transpiration. Thus it will be evident that unless the air dew-point temperatures are less than the snow surface temperatures (vapor pressure gradient directed upwards) there can be no evaporation. If, as may frequently be the case, the air dew-point temperatures exceed that of the snow cover (vapor pressure gradient directed downwards) there must be a frost condensation occurring on the snow surface. This explains what took place between the 16th and 20th inclusive when an appreciable amount of frost was condensed upon the snow surface.

On the 22d and 23d only patches of snow remained on the field and by the morning of the 24th it had entirely disappeared. The increase of evaporation during the period between the 22d and 27th inclusive over that of the period 14th to 21st, inclusive, is to be attributed directly to the disappearance of the snow cover. However, even with the disappearance of the snow it is to be noticed that the average daily evaporation rates for the period from the 21st to the 27th, inclusive, were lower than those for the period prior to the first snowfall. This is accounted for by the greatly decreased temperature of the ground during the last half of the month.

Cold and dry Polar Continental air occupied the area on the 27th but was displaced by a southwesterly current of transitional tropical air on the 28th. The dew-point temperatures of the polar air averaged approximately 12° but increased to about 20° in the modified air. The ground temperatures were still quite low and with the invasion of the modified air with increased specific humidities, the moisture gradient was directed downward for a good portion of the day and appreciable condensation of frost occurred. The 29th was overcast and light rain began at 11 a. m. and continued through midnight. The cloud base which remained below 1,000 feet lowered to the ground in the evening, resulting in foggy conditions. On the 30th a low stratus overcast with ceilings varying between 200 and 600 feet and light rain or mist and fog persisted during the entire day. The rain ceased in the early morning of the 31st when a polar air mass again invaded eastern United States. Practically no evaporation occurred on the 29th or the 30th. On both of these days, however, significant amounts of condensation occurred. The small evaporation of the 31st was associated with the invasion of the polar air mass. The total amount of evaporation at the Arlington Farm for the month of January 1939 amounted to 0.4747 inches. The total amount of condensation as computed was 0.0632 inches. It is recognized that this latter figure may be somewhat exaggerated.

In general, most researches on the problem of evaporation have been directed toward the determination of moisture losses from bodies of water. This has led to a deficiency in basic and quantitative information regarding actual moisture losses from watersheds or land surfaces. Studies of the evaporation from free water surfaces alone have proved wholly inadequate from the standpoint of soil conservation. What is required is the actual moisture losses from all types of geographic surfaces, or in other words, a quantitative estimate of the evaporative phase of the hydrologic cycle. With this information the soil conservationist will then be able to interpret the interrelations between climatic forces and land utilization (**36**).

In an effort to supply this much-needed information a method for determinining the evaporation from either land or water areas has been presented in this preliminary report. The practicability of the technique has been completely demonstrated. It is hoped that with proper instrumental installation it will be possible to determine transpiration rates and moisture requirements of various field crops and forest trees, the effectiveness of various moisture conserving practices, and the relative importance of evaporation and transpiration in the hydrologic cycle. However, the question of satisfactory instrumentation has not been entirely solved and the validity of various theoretical assumptions has yet to be tested. These questions are expected to be discussed further in a detailed report on the problem of evaporation to be published in the near future.

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VALLEY-HEAD CLOUD WINDOWS

By Ronald L. Ives

[Boulder, Colo., September 1938]

"Windows," which hold their positions, regardless of cloud motion, have been observed in the cloud strata over the heads of many glaciated valleys in the Colorado Front Range area.

Observations during the summer of 1938, when abnormal wetness in the foothills and on the adjacent plains increased the cloudiness on the east flank of the Front Range and broke up the normally present "convection sheets" into valley winds, indicate that the cloud windows are caused by these valley winds.

Following the courses of the valleys upward from the plains, the winds have a temperature at 12,000 feet of about 75°. Mountain air, at the same time and elevation, has a temperature of about 50°, and has a greater humidity than the valley air.

Cloud strata are at 12,500 to 13,000 feet and 14,000 to 16,000 feet: The lower stratum being confined closely to the range; the upper stratum covering the parks between the ranges, and sometimes extending several miles over the plains.

Valley winds, on reaching the cirques at the valley heads, are diverted sharply upward, and the clouds are evaporated as the winds rise through them. Clouds that drift into these updrafts are immediately dissipated. No cloud

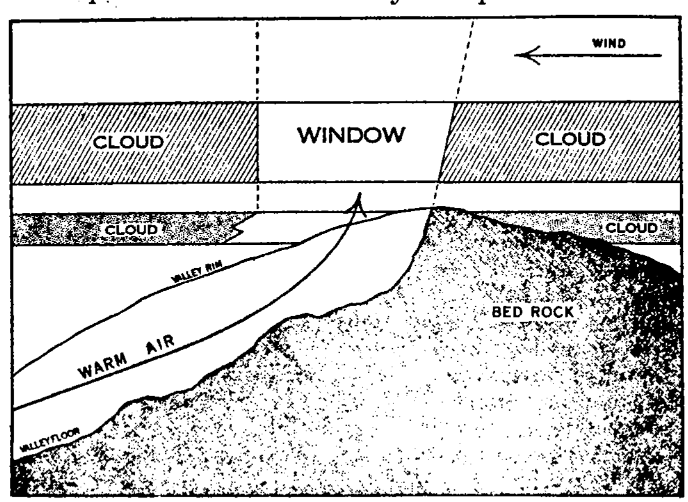

FIGURE 1.—Typical valley-head cloud window.